JGR - Atmospheres Apr 2001 (in press)

Effects of Glacial Meltwater in the GISS Coupled Atmosphere-Ocean Model: Part I: North Atlantic Deep Water Response

David Rind, [1] P. Demenocal, [2] Gary L. Russell, [1] Sukeshi Sheth, [1] Dan Collins, [1,3] Gavin A. Schmidt [1] and J. Teller [4]

Abstract

Varying magnitudes of freshwater input through the St. Lawrence are added to different versions of the GISS coupled atmosphere-ocean model. The results show a generally linear response in percentage North Atlantic Deep Water (NADW) reduction to the volume of water added without any obvious threshold effects. When the estimated best-guess freshwater input for the Allerod-Younger Dryas interval is added, only small reductions in NADW production occur in less than a century, with complete cessation requiring more than 300 years. If the ocean circulation was weaker in the Allerod preceding the Younger Dryas, somewhat larger percentage reductions occur, although it would still take 150–200 years for a complete shutdown. The freshening is aided by a positive feedback associated with the reduction in evaporation relative to precipitation in the North Atlantic, amounting to some 20–30% of the added freshwater runoff. When the freshwater input is stopped after complete NADW shutdown, NADW production does not resume. With complete NADW cessation, cooling eventually occurs throughout the Northern Hemisphere, including regions where it is not thought to have arisen during the Younger Dryas, and the globe cools by 1°C due primarily to the sea ice albedo feedback; while in many regions the observed cooling is reproduced, the cooling in Greenland is much less than is estimated. With only 50% NADW cessation, cooling is primarily circum-North Atlantic. No rapid warming is seen in response to freshwater cessation. The sensitivity displayed here for NADW reduction due to freshwater input from the St. Lawrence is roughly similar to that associated with increased precipitation accompanying global warming in the next century, so a similar sensitivity may arise with Younger Dryas freshwater inputs from other suspected locations (Greenland, Europe).

I. INTRODUCTION

Considerable indirect evidence exists that North Atlantic Deep Water (NADW) production has oscillated during the glacial age, and perhaps during the Holocene as well (Bond et al, 1997), primarily from benthic foraminiferal δ^{13}C (e.g., Curry et al., 1988, Zahn et al., 1997), and Cd/Ca data ((e.g., Boyle and Keigwin, 1987; Keigwin et al., 1991). Paleo-indicators of surface ocean temperatures, including planktonic foraminiferal and diatom assemblages (e.g., Bond et al., 1993; Karpuz and Jansen, 1992), alkenone (Sachs and Lehman., 1999) and δ^{18}O data (e.g., Cacho et al., 1999) suggest colder conditions along with these oscillations, which in general match the oscillations in δ^{18}O in the Greenland Ice Cores (the "Dansgaard-Oeschger [D-O] events") (e.g., Johnsen et al., 1995). The occurrence of ice-rafted debris [IRD] in Heinrich Events [HE], are thought to represent millennial-scale cooling extrema which are followed by abrupt warming.

Reductions in NADW have been hypothesized to be associated with freshwater inflow to the North Atlantic, either from glacial melting (e.g., Broecker et al., 1985), or from ice sheet instabilities leading to the influx of fresh water from IRD (e.g., Alley and MacAyeal, 1994). Several studies have documented large reductions in NADW production associated with the millennial-scale glacial cooling events (Curry et al., 1999; Zahn et al., 1997; Charles et al., 1996). Records from the Feni Drift at 12°W near Ireland do suggest a close match in time between NADW changes as deduced from 13C shifts and the Younger Dryas interval as deduced from faunal assemblages and IRD changes (Bond et al., 1997). To the extent that the colder time periods are associated with reduced NADW production, the likely concurrent reduction in transport of heat by ocean currents is thought to be a prime factor. With reduced heat transport, colder conditions allow the glaciers to stabilize, reducing the freshwater input; in turn, the reduced NADW outflow of salinity from the North Atlantic leads to a salinity build-up and reinitiation of deep water production, with subsequent warming (Broecker et al., 1990). This scenario has two distinct parts, each of which can be subjected to observational or modeling scrutiny. (1) Did fresh water input lead to rapid reductions in NADW? (2) Did fluctuations in NADW production lead to the cooling observed in marine and land paleo-temperature indicators? With respect to question (1), significant doubts have arisen, associated with a mismatch in the timing of outflow from the Laurentide Ice Sheet relative to NADW changes, and the associated lack of the expected salinity signal in the Gulf of St. Lawrence (Keigwin et al., 1991; Fairbanks, 1989; de Vernal et al., 1996; Moore et al., 2000). Modeling studies with ocean GCMs or simpler models imply a sensitive, rapid response to salinity changes (e.g., Weaver and Hughes, 1994; Rahmstorf 1995). However, in simulations with a coupled atmosphere-ocean model utilizing freshwater inputs on the order of 0.1 Sv, similar to observed estimates from either glacial meltwater (Licciardi et al., 1999) or ice rafted debris (MacAyeal, 1993), NADW response is gradual, reaching a complete shutdown in some 500 years (Manabe and Stouffer, 1997; 2000), much longer than the initiation time of many of the observed oscillations. Schiller et al. (1997) using a relatively large fresh water input (up to 0.6 Sv) did find a complete deep water cessation in several hundred years. Fanning and Weaver (1997a) also calculated that if the North Atlantic had been preconditioned by glacial waters from the Mississippi, NADW could be shut down within 200 years, a time period consistent with some observations (e.g., Hughen et al., 2000). Fresh water input is unlikely to have been maintained over such a long period, occurring in much shorter pulses (Moore et al., 2000). A related observation that spans both questions is that the lowest salinity is found when warming switches on at 13.4kBP (all dates are in calendar years), rather than having increased sufficiently to restore NADW production and the heat transports leading to the warming at that time (Karpuz and Jansen, 1992).

There are further inconsistencies associated with the second question, as some studies have concluded that the observed rapid temperature changes did not coincide with the termination or initiation of NADW production (Zahn et al., 1997; Moore et al., 2000), with additional evidence for a surface cooling precursor event prior to maximum IRD and NADW reductions associated with each Heinrich Event (Bond and Lotti, 1995; Bond et al., 1999).

Some of these issues are addressed in a set of coupled atmosphere-ocean model simulations with the GISS GCM (Russell et al., 1995). When run for assessments of climate change during the next century associated with greenhouse gas warming, this model produces a reduction in NADW due to freshening of the North Atlantic, with added moisture transport from the warming tropics (Russell and Rind, 1999; Russell et al. 2000). How it responds to freshwater inputs from North America is the subject of this paper. The response of Antarctic Bottom Water, and the Southern Hemisphere in general, is the subject of Rind et al. 2001 (henceforth Part II).

II. MODELING EXPERIMENTS

The coupled atmosphere-ocean model used here has a horizontal resolution of 4°x 5°, with 9 levels in the atmosphere and up to 13 levels in the ocean (Russell et al, 1995, 2000). While this is coarse resolution considering the fine-scale nature of convective deep water events, there are several features of the model that provide for higher resolution. First, convection is calculated on quarter-box resolution. In addition, effectively finer resolution is obtained for heat and salt advection through the use of a linear upstream scheme which keeps track of the first order moments in addition to mean values. As shown below, the model's NADW mass streamfunction is in good agreement with observations for the time period of the experiments, although it is slowly decreasing with time. Ultimately, all such experiments should be performed with as high resolution as practical; for Parts I and II of this study, the coupled atmosphere-ocean model was used for simulations of more than 1400 years.

Several model features are optimal for these experiments. The model has a free surface, so fresh water can be added directly, rather than parameterized as a salinity change as has been done in previous experiments of this type with coupled models. River flow is included in the model, and the freshwater input can be added directly to the river mouth, rather than being input uniformly throughout a latitude zone in the North Atlantic, as has generally been done. The temperature of the incoming meltwater is set to the freezing point of fresh water (although in reality it may have been as warm as 2–5°C).

Multiple experiments with sustained gradual freshwater input are performed primarily with three different versions of the model. The standard version (henceforth called STAND) has generally realistic NADW production without the use of flux corrections, at least over the length of time of the experiments conducted here (200 years) (Russell et al., 2000). (A slightly different version of this model, begun as an extension of STAND after 90 years, and having a weaker North Atlantic stream function, will be referred to as STAND (EXT); it is used for a particular experiment described below.) We also use the previous version of the model, whose NADW production was weak, somewhat shallower, and further south (WEAK). The primary difference between these two versions that accounts for the change in NADW strength is the replacement of the original vertical mixing and convection scheme with the k-profile parameterization (KPP) vertical mixing scheme of Large et al. (1994) in the later model. While the weaker circulation is not an exact analogy to the ice age circulation, the changes are in the direction of what was expected to prevail during the glacial time frame, in general. For the third version, we add salinity flux corrections (FLUX) to the older model (STAND), which increases the NADW production. This experiment is analogous to those performed elsewhere (e.g., Manabe and Stouffer, 1997, 2000), and was done in an attempt to understand whether the response differs depending upon whether flux corrections are used, as suggested by experiments testing potential instabilities in flux corrected models (Tziperman et al. 1994). The flux adjustment procedure differs from that of Manabe and Stouffer primarily in that only freshwater adjustments are made (not heat fluxes as well), to understand how they affect results associated with freshwater (i.e., salinity) forcing. A complete listing of the control runs and experiments is provided in Table 1.

For each model we alter the freshwater input from the St. Lawrence by factors of 2,4, and 8 over its instantaneous values in the run, with the changes put in continually over the course of the experiments. The freshwater is added at 0°C into the first three levels of the ocean, similar to how the river flow is normally added in the model. The freshwater fluxes are approximately 0.03 Sv, 0.06 Sv, 0.12 Sv in the experiments, respectively, with only small changes from one model to the other. Given that the instantaneous flow can differ in the experiments as a function of time, so does the actual value added, but in each decade the fluctuations are less than 10% of the mean value.

Two additional experiments were conducted with STAND. In the first, we add 32 times the instantaneous freshwater input to the North Atlantic, amounting to about 0.53 Sv. In the second, we use the more realistic estimates of freshwater input derived by Licciardi et al., 1999) [called LTC(1)] which includes precipitation plus ice melt; it also includes freshwater fluxes into a variety of locations in addition to the St. Lawrence. The latter experiment was also repeated with STAND (EXT) (called LTC(2)) in an attempt to gauge the variability of the results. The freshwater flow into the Atlantic through the St. Lawrence in the LTC experiments is about 0.15 Sv for the time period 13–11.44 cal ka (Licciardi et al., 1999, Appendix A), and there was an approximately equivalent amount of fresh water coming into the Atlantic from other locations, the chief contributor being the Hudson Strait. Overall, these varying flux experiments allow us to determine this model's North Atlantic sensitivity to salinity perturbations (e.g. Weaver and Hughes, 1994).

Each control and experiment was run for at least 160 years (except for FLUX, which was run for 120 years), starting from a 23 year spin-up from observed initial conditions (Levitus 1994), the same procedure as has been employed for simulations of the next century (Russell and Rind, 1999, Russell et al. 2000). (The exception is LTC (2), which started from a control run simulation with a 90 year spin-up). It is not known how the results would differ were the models to start from a coupled equilibrium state, which would not only require unavailable amounts of computing time but would also likely result in greater deviations from the observed ocean-atmosphere system.

As shown in the next section, even the control run with the KPP scheme (and no flux corrections) (STAND) shows a drift in the magnitude of NADW circulation, with a reduction on the order of 20% during the first 160 years. While this is much smaller than the reductions in NADW in most of the freshwater input experiments, it may imply a tendency for the model to suffer reductions that would not take place in a completely stable model. The version with flux corrections (FLUX) which is completely stable, although at a smaller value of the streamfunction, actually shows a greater sensitivity to freshwater input, so if stability is important, it is not more important than other factors (strength of the circulation, flux corrections, etc.).

There was also no attempt to use 13 cal ka conditions over land, an effect which might change the results. The WEAK control run had greater sea ice in the North Atlantic and a more southerly Gulf Stream, hence somewhat closer to the ice age configuration, which may be important in translating climate change influences equatorward (Manabe and Stouffer, 2000). At the end of the freshwater simulation experiments, in the 32× and LTC(1) input simulations with STAND we allowed the model to continue without additional freshwater fluxes, to determine whether it would return to its original state within a short period of time, and whether any oscillations were induced in the process.

The duration of the runs fits within the general concept of the time it took for the Younger Dryas Oscillation to be initiated (on the order of 100 years at most - D. Peteet, personal communication), and terminated, on the order of 10 years (Karpuz and Jansen, 1992; Alley et al., 1993).

III. RESULTS

A. North Atlantic Deep Water Response

Shown in Figure 1 is the streamfunction for the North Atlantic at 50°N, 900m in the different control runs and the 8× St. Lawrence experiments. Over the 160 years shown, the STAND control run streamfunction (Figure 1, top) decreases by about 20% from its value in the first two decades, although it is relatively steady for the last 120 years at a value close to the recent observations of 17 Sv derived from chlorofluorocarbon inventories (Smethie and Fine, 2001). Whether this degree of instability affects the model sensitivity is not known. In comparison, the streamfunction in WEAK (Figure 1, bottom) decreases to small values at this latitude (it is somewhat stronger further south). Greater stability occurs in FLUX with salinity corrections, although it is stabilized at a value between that of WEAK and STAND.

For each control run, the addition of 8× St. Lawrence fresh water, about 0.12 Sv, results in a weaker stream function, with the absolute reduction increasing with time (except toward the end of WEAK where the control run values became quite small). With this amount of freshwater input, no complete cessation of NADW production occurs within ≈150 years.

The percentage change induced in the North Atlantic stream function at 52°N, 900m in depth is given in Table 2 (the percentage change allows for some uniformity in comparing the decreases, considering the different control run magnitudes; actual decreases in Sv can be determined by using Table 2 in conjunction with Figure 1). In the standard experiment, freshwater fluxes on the order of 0.03 Sv have little effect for the first 100 years but eventually result in a small reduction (11%). With fluxes on the order of 0.06 Sv, a small effect occurs by year 75 and increases later in the run. With freshwater inflow of 0.12 Sv, a 10% reduction occurs by year 25, and slowly grows with time. With inflow of some 0.5 Sv an effect arises by year 10, and the effect grows rapidly, with essentially complete deepwater cessation at this latitude by the end of a century (reduction greater than 100% indicates a change in sign of the streamfunction). Both the rapidity of the effect, and the magnitude of the response, increase with increasing freshwater input.

The relationship between the stream function change in percent (y) in response to the fresh water input is given in Tables 3 (a-c). In Table 3a, the results are related to freshwater input (x) in Sv, and are shown for different time periods. In Table 3b, the results are relative to time (t) in years, and are shown for different freshwater input magnitudes. In Table 3c the results are relative to Volume (Sv-yrs). As the tables show, greater NADW percentage reduction is associated with greater magnitude of input, or with greater duration, hence with greater volume input. The results are highly linear; the linear correlation coefficients shown are often greater than 90%. There is no "cliffhanger" effect in which deepwater is suddenly diminished.

The tables also show that the system appears to be more sensitive with the weaker streamfunction (WEAK), and least sensitive with the strongest streamfunction (in STAND), in agreement with previous modeling results (Tziperman, 1997; Fanning and Weaver, 1997b). The results for FLUX are intermediate, as, in general, are its streamfunction values. The results when the realistic fresh water was added in STAND (and STAND (EXT)) are also shown in Table 2. Utilizing the formulae developed in Tables 3(a-c) for that experiment would predict that the input through the St. Lawrence of about 0.15 Sv for 100 years would result in about a 25% reduction, slightly more than what occurred on average in LTC (1) and similar to that in LTC (2). It does not appear that the additional fresh water being put in elsewhere reduced the stream function any further than what would be expected from its St. Lawrence input, a result generally consistent with that of Manabe and Stouffer (1997; 2000), who showed that in their model, freshwater added far from the NADW source (in their case, near the Caribbean Sea), was less effective in producing an NADW response. The greater stream function reduction in LTC (2) after 70 years is consistent with the general conclusion that a weaker North Atlantic circulation is more susceptible to reduction by freshwater input. What the more diffuse source did do was make the streamfunction decrease more uniform throughout the North Atlantic, for in the experiments with moderate increases in St. Lawrence river input only, the reduction in streamfunction at 28°N were generally less than those at 52°N. In the extreme freshwater experiment (32× St. Lawrence), the stream function value eventually decreased to zero at 28°N as well.

B. Salinity Response

The change in NADW production is related to the salinity changes induced by the added fresh water input. Shown in Figure 2 are the salinity changes in STAND (and STAND (EXT)) for the different experiments during years 101–120, as well as the values for the 8× St. Lawrence increase in WEAK and FLUX. Largest freshening is found near the St. Lawrence, with an effect that is carried downstream to the northeast, and increases with increasing freshwater input. Some difference can be seen among the runs, with WEAK and FLUX showing larger salinity changes for the 8× St. Lawrence input, consistent with their greater percentage changes in stream function (and the reduced salinity transport). The difference between LTC (1) and LTC (2) is also consistent with the greater reduction in the latter experiment. With a factor of 32× increase in the St. Lawrence freshwater input, salinity reductions occur throughout the North Atlantic, and even somewhat south of the equator.

The effect is amplified somewhat by the resulting climate response. Shown in Figure 3 are the changes in precipitation and evaporation in the 32× St. Lawrence run and the LTC(1). In both runs, but more so in the former, evaporation decreases associated with the colder temperatures are in general larger than the precipitation decreases, a response that further decreases the salinity. This positive feedback was not included in the "salt oscillator" theory of Broecker et al. (1990) and would, in conjunction with freshwater input into the North Atlantic, presumably act to reduce the ocean circulation.

This latter effect is not small. Over an area in the North Atlantic from 38°–70°N, and 0–70°W, or about 20x106 km2, in the control run evaporation exceeds precipitation by 2.11 mm d^{-1}, while in the 32× St. Lawrence run, evaporation exceeds precipitation by only 1.41 mm d^{-1}. The difference, 0. 70 mm d^{-1} over this area amounts to a freshening of 0.16 Sv - or about 30% of the input freshwater through the St. Lawrence in that experiment. Hence the actual reduction in NADW is being produced by salinity forcing 1.3 times the value used as input in the experiment. In the case of LTC (1), evaporation exceeds precipitation by 1.98 mm d^{-1}, so the difference 0.13 mm d^{-1} or 0.03 Sv is 22% of the runoff through the St. Lawrence. That the result percentage-wise is almost the same in the two experiments may help account for the linearity of the percentage change in NADW production.

The salinity change for various decades is shown in Figure 4 for the 32× and LTC(1) experiment. The salinity decrease in the North Atlantic is continually increasing with time, and eventually occur in the Pacific as well (where as shown in Figure 3 the evaporation decrease exceeds the precipitation decrease). Salinity changes are also occurring in the Southern Hemisphere, an effect which will be discussed further in Part II.

C. Temperature Response

The temperature changes for the different experiments for years 101–120 are given in Figure 5. Cooling with increasing magnitude arises in the North Atlantic region, with more general Northern Hemisphere cooling with increasing fluxes. Cooling is more intensive at high latitudes in WEAK, again associated with the greater NADW percentage reduction. With the LTC freshwater input, cooling is primarily at high Northern Latitudes, somewhat greater in LTC(2). The tropics in general experience little temperature change or slight warming, while at high southern latitudes, there is a tendency for cooling with more extreme fluxes or weaker control runs.

The temperature change for different time slices is shown in Figure 6. By the end of the freshwater input simulation most regions of the globe have cooled in the 32× input run, whereas with the LTC input the cooling in LTC(1) is confined primarily to the high northern latitudes, especially over the North Atlantic and downstream.

To understand the difference between the results with the excessive freshwater input (32×) compared with the realistic input (LTC(1)), shown in Figure 7 are the ocean/sea ice changes as a function of time for the Norwegian Sea. The greater freshwater fluxes in the 32× St. Lawrence run quickly result in reduced ocean heat transport convergences, due to the reduction in North Atlantic Deep Water production. The cumulative effect of this heat transport reduction is to generate colder sea surface temperatures, but, given the heat capacity of the Norwegian Sea region, it takes about 60 years for the temperature changes to arise. An additional factor adding to the cooling is the growth in sea ice cover, which begins about year 80. By increasing the surface albedo it deprives the ocean of solar energy absorption, enhancing the cooling. In comparison, the LTC input results in small changes in NADW production, ocean heat transport convergences, and sea surface temperatures, with little in the way of sea ice response. For the 8× St. Lawrence input of slightly more than 0.1 S, only WEAK produced cooling, of 2 to 3°C associated with reduced heat flux convergences and increased sea ice (not shown), and the effect actually diminished after year 150.

With respect to the surface air temperature changes over Greenland (Figure 8), cooling at the end of 150 years in the 32× run reaches some 7°C at a time when NADW production is essentially zero in the Northern North Atlantic. Manabe and Stouffer (2000) found strong oscillations, of some 6°C in sea surface temperature within about 50 years. The model here shows somewhat smaller oscillation effects, with initial warming during the first few decades, the sea surface temperature perturbations being generally less than 1°C (Fig. 7), and the temperature fluctuations over Greenland being on the order of 2°C. We return to this point in the discussion section. With the apparently more realistic LTC input, there is only slight cooling over Greenland through the length of the simulation.

D. Ocean Transport Response

Shown in Figure 9 are the northward energy transports in the Atlantic during the last two decades of the 32× St. Lawrence fresh water input to STAND. As expected, poleward energy transport in the North Atlantic has declined dramatically. Poleward energy transports in the South Atlantic have increased associated with an increase in Antarctic Bottom Water Production, as discussed in Part II. The geographical distribution of the change in heat (enthalpy) transport between the 32× St. Lawrence experiment and the control for years 141–160 is given in Figure 10. Reduced poleward transport is visible in the western North Atlantic as expected with NADW production eliminated. The surface ocean current changes are given in Figure 11. The reduction in NADW reduces the deep southward flow of water out of the Atlantic, and so also reduces the northward flow of water (and energy) at upper levels into the Atlantic, visible as a reduction in the Northeast Brazil current, the Florida current, and the Gulf Stream. An intensified eastern Greenland current arises in response to the reduced mass flow of the North Atlantic drift, and the anomaly in the ocean current direction is such as to allow for ice rafted debris to be advected from the Labrador/Greenland area southeastward in the Atlantic, as observed during Heinrich events (e.g., Bond et al., 1993).

However, in regions of the northern Atlantic poleward transport actually increases. This is due at least in part to heat transports associated with the wind driven circulation. It will be shown in Part II that high pressure develops over the North Atlantic in conjunction with the oceanic cooling. The anticyclonic circulation around the high provides for more poleward flow over the central North Atlantic, as can be seen by the change in ocean currents in the first model layer (Figure 11). Overall, the northward heat transport circa 60°N in the Atlantic is reduced by only about 50% (Figure 9) as are ocean heat transport convergences poleward of 60°N, where values decrease from $30\mathrm{Wm}^{-}2$ to $15\mathrm{Wm}^{-}2$. The maintenance, and indeed strengthening of the wind driven circulation has helped limit the cooling over Greenland (Figure 8). At other locations, (e.g., the North Pacific), the currents respond to changes in atmospheric forcing, hence displaying a more anticyclonic circulation where sea level pressure increases. The results indicated here occur to a lesser extent in the other runs and at shorter time periods.

E. Global Radiative Response

To understand the global temperature responses, we show in Table 4 the relevant radiative characteristics for the 32× St. Lawrence simulation during years 141–160 relative to the control run. The changes in the other simulations and time period were of a similar nature, but reduced in magnitude.

At this stage in the extreme freshwater input experiment, the global surface air temperature has decreased by a little more than 1°C; it will be shown in Part II that the change is not noticeably greater with more extended NADW shutdown. The effect is primarily in the Northern Hemisphere, although a small cooling is occurring in the Southern Hemisphere. The vertically-integrated atmospheric air temperature change is not as large; the surface cooling being primarily at high northern latitudes remains near the ground due to the large atmospheric stability in that region.

Overall, atmospheric water vapor decreases by 4–5%. Given the magnitude of the cooling, this is a little more than 1/2 the water vapor response that occurs in 2×CO2 experiments (assuming linearity) , where a 4°warming produces a 30% water vapor increase. The reduced water vapor change is undoubtedly due to the lack of response in the tropics.

Sea ice cover increases, especially in the Northern Hemisphere, and that leads to a surface albedo increase, again primarily in the Northern Hemisphere. Total cloud cover decreases slightly, and the effect of clouds and cloud changes is to mute the planetary albedo change by about a factor of 2 relative to the surface albedo response in the Northern Hemisphere and globally, although an increase still occurs.

The planetary albedo increase implies a net radiative loss to the system of 1.16 $\mathrm{Wm}^{-}2$. Comparing that result to the surface air temperature change, the model is responding with a sensitivity of about $1°\mathrm{C/Wm}^{-}2$, in this transient change experiment. For a corresponding increase of CO2 at 1%/year, with a $4\mathrm{Wm}^{-}2$ forcing at the time of doubled atmospheric CO2, the warming was about 2.5°C, for a sensitivity of $0.625°\mathrm{C/Wm}^{-}2$. Neither experiment is in equilibrium with the radiative forcing change, and both have been experiencing forcing for over 100 years, although the forcing has remained constant in the freshwater input experiment, while it was increasing with time in the increased CO2 run. This difference in itself may account for some if not all of the different sensitivities. To the extent that the greater sensitivity in the freshwater input experiment is real, it could result from at least two factors: the cooling being primarily at high northern latitudes where the temperature change is kept near the surface; and the reduction in NADW circulation implying less mixing of the response to depth in the ocean (the increased CO2 experiment also had a reduction in NADW production, but only by about 30%).

The last line of the table shows the net radiation at the top of the atmosphere during this time period. The small positive value implies more energy is coming in than is leaving, and hence the atmosphere would warm. If it followed the same sensitivity as that indicated above, warming of a few tenths °C would be expected. The positive radiation imbalance is totally in the Southern Hemisphere, which would then likely warm sufficiently to offset the hemispheric cooling indicated in Table 4.

F. Response to Freshwater Cessation

In both the 32× St. Lawrence and the LTC(1) experiments, simulations were continued after year 160, but now the additional freshwater input was ended. In previous experiments with freshwater input ended, Manabe and Stouffer (2000) found that NADW started recovering almost immediately, as did Schiller et al. (1997), although it took about 120 years for full recovery. In these experiments, the question was whether a rapid recovery would occur, relevant to issues of rapid temperature warming at the end of the Younger Dryas. The relevant results are shown in Figures 7 and 8 as a continuation past year 160. In the 40 years through year 200, conditions in the Norwegian Sea (Fig. 6) show a small amelioration of effects, with SSTs warming slightly, ocean ice cover decreasing slightly, and converged ocean heat transports increasing somewhat, although the latter result shows much variation with time. Air temperatures in Greenland (Fig. 8) warm a little, but with no obvious continuing trend. NADW production showed no increase through year 200. Insofar as a rapid response is concerned, as has been observed for the warming in Greenland from ice core isotopic reconstructions, ending the freshwater input did not produce that result. Whether a recovery would occur over a longer time period is discussed in Part II.

IV. DISCUSSION

In the introduction, we raised two separate questions: did freshwater input lead to rapid reductions in NADW, and did changes in NADW production lead to the cooling observed during events such as the Younger Dryas? In this section we deal with each question separately, and briefly address the issue of freshwater input to the North Atlantic in general.

A. Did freshwater input reduce NADW rapidly in the Younger Dryas?

As indicated by these results, adding freshwater to the North Atlantic through the St. Lawrence does result in a decrease in NADW production. The effect is not rapid with realistic freshwater inputs, for which NADW we estimate (using the volume/stream function change relationship) that NADW cessation would take some 350 years to occur. To reduce NADW production to negligible values in the northern North Atlantic within the 200 year time frame indicated by some observations would require St. Lawrence freshwater input about 1.5 to 2× that estimated to have occurred (if the circulation was initially as strong as it is today). Weaker freshwater inputs produce a proportionaly weaker percentage change in NADW production. As discussed in Part II, the Atlantic AABW changes are of the opposite sign to NADW changes, which would lead to an overestimate of NADW changes in glacial times when using any water mass tracer including Cd/Ca and d13C.

Input in regions outside of the St. Lawrence does not appear to hasten NADW reduction, although all regions have not been tried (western Europe, for example). While the control run in STAND is itself showing a slow decrease in NADW production with time, if that acts to make it more sensitive to freshwater forcing, then even less of a response would be found in a completely stable model.

A primary result from these experiments is that NADW circulation responds percentage-wise in a linear fashion to freshwater volume input, hence linearly as a function of input freshwater magnitude, or input freshwater duration. A linear response was found over a small range of freshwater forcing (¡0.1 Sv) in the global ocean circulation model experiments of Rahmstorf (1995); the major difference here is that they extend to much larger freshwater inputs (0.5 Sv, at least), despite obvious nonlinearities in the salinity response. Not every response shown in Table 2 is linear; variations occur from decade to decade, as they do in the climate system. Yet the overall high correlation coefficients of NADW percentage reduction to freshwater input shown in Table 3 argues against the threshold effects and resulting nonlinearity discussed by Rahmstorf (see also Rahmstorf, 2000). Schiller et al. (1997) also found a "rather smooth" effect of meltwater input on formation rate of NADW, and our results agree with their conclusion about the absence of highly nonlinear responses. Evaporation minus precipitation becomes more negative as NADW (and its associated heat transport) changes, adding to the freshwater dilution effect; the fact that the addition was proportional to the freshwater input may have helped maintain the linearity in NADW response. This linearity exists in all three models used for these experiments, and hence does not depend on the magnitude of the control run circulation, or the presence of salinity flux corrections, although the magnitude of their individual responses differs.

Our results compare favorably with previous experiments that have resulted in complete NADW shutdown in simulations requiring hundreds of years, often with exaggerated freshwater input. There is little evidence from these experiments that rapid, large-scale temperature fluctuations would have arisen with freshwater input of several tenths of a Sv, although an eventual large response to such a forcing is likely. Observations that NADW production decreased at a time significantly prior to the onset of the Younger Dryas might therefore be appropriate (Zahn et al., 1997), as might be the cumulative nature of the D-O events in a "Bond" cycle prior to a Heinrich event, which is thought to possibly represent a complete NADW shutdown. The large, rapid oscillations seen in the model runs of Manabe and Stouffer (2000) do no t arise in these experiments, which feature smaller oscillations. In these experiments freshwater is literally being added, rather than salinity being abruptly changed while mass is kept constant, although it is not known whether that is the reason for the difference. Just as realistic freshwater input did not produce a rapid reduction in NADW production, neither did ending the freshwater input result in its rapid reinvigoration. In the original conception of Broecker et al., (1985), colder conditions during the Younger Dryas (or any D-O oscillation) would lead to reduced glacial meltwater runoff, allowing salinity to build-up in the North Atlantic. Hence when freshwater fluxes were reduced, NADW production would re-emerge. When the freshwater fluxes were ended in these experiments (and those discussed in Part II), there was no indication of any rapid return of NADW production or large, rapid warming; the only warming that took place appeared to result from the cessation of cold water additions at 0°C and resultant sea ice decrease. This conclusion agrees with the observation that the termination of the Younger Dryas, or any of the abrupt events, did not appear to coincide with the return of NADW production (Zahn et al., 1997; Moore et al., 2000). The more extreme experiments here are closer to the two stable equilibria found by Manabe and Stouffer (1988), in which once a complete NADW cessation occurred, in response to an input of 1 Sv of fresh water, deep water production never recovered. In contrast, in Manabe and Stouffer (2000), with an input of 0.1 Sv, a weak, shallow circulation cell was maintained, which then did recover when freshwater input ended. Schiller et al. (1997), with an input of 0.6 Sv, also found complete NADW shutoff, but when the freshwater input was abruptly ended, deep water fully recovered by 120 years. Manabe and Stouffer (2000) speculated that a vertical diapycnal diffusion coefficient as low as a few tenths cm^2s^-1 would enable a coupled model to maintain the "no deep water production" mode, in contrast to what they assumed was the higher diffusion used in the Schiller et al. (1997) study. In this model, the explicit vertical diffusion is 0.3 cm^2s^-1, while the linear upstream scheme used for heat advection is not strongly diffusive. Deep water production was not reinitiated, in line with the Manabe and Stouffer suggestion and the studies of Rahmstorf (1995).

The version of the model with a weaker NADW circulation, and more extensive sea ice, was more sensitive to freshwater input, in terms of its percentage change in NADW, and its corresponding cooling. This version might be more appropriate for some time periods during the glacial cycle, although the situation for the Allerod preceding the Younger Dryas is somewhat uncertain. Ruddiman and McIntyre (1981), from the position of the polar front, concluded that NADW production was back to current day values, and Moore et al. (2000) reviews evidence for this conclusion, which is applicable in the region south and east of Iceland. However, north of Ireland in the Greenland Sea the Allerod was cold, and in the Denmark Strait and the eastern Nordic Seas, the Allerod was not as warm as currently. For the version of the model with the weaker streamfunction (WEAK), we estimate from Table 3c that it would take some 150 years for a freshwater input of 0.15 Sv to produce a 100% decrease in NADW production. It is not known what influence glacial land ice boundary conditions would have on the sensitivity of NADW to these freshwater inputs. The greater resistance to fresh water input of a Holocene-type circulation (in STAND) might help explain the apparent weak response to the last meltwater pulse, circa 10 kyr BP calendar years (Karpuz and Jansen, 1992), when NADW strength may have already been back to current values.

B. Did changes in NADW production result in the cooling of the Younger Dryas?

If present estimates of the location of cooling associated with the Younger Dryas are correct, these model simulations cannot produce all the observed magnitudes and geographic pattern of response. Using the results shown in Rind et al. (1986) and Peteet (1995), the Younger Dryas cooling was most apparent in the circum North Atlantic over land and in the eastern North Atlantic in the ocean. As seen in Figure 6, this is approximately true for years 41–60, but in the later years with complete NADW suppression, extensive cooling on the order of -1°C is found throughout the Northern Hemisphere, including over central North America. Shane et al. (1993) did report a cooling of similar magnitude in Ohio but most other areas of the continental United States show no such effect, either because no such cooling happened (Rind et al., 1986) or palynological gradients were too small to detect such a change. Cooling of the hemisphere as a whole results from the increased sea ice and planetary albedo, with water vapor feedback as given in Table 4. [Interestingly, the hemispheric-wide expression did not occur to the same extent in the coupled model simulations of Manabe and Stouffer (1997), perhaps because cold freshwater was not literally being added. We return to this question in Part II.] To keep the cooling primarily circum-North Atlantic, a NADW reduction on the order of 50% is a better match then a more extensive NADW change. On the other hand, more extensive NADW change is required to produce cooling in the Santa Barbara basin, as apparently occurred (Hendy and Kennett, 2000), and intensification of the winter monsoon (as indicated by the temperature change in southern Asia).

There is also some question as to whether the magnitude of cooling in the Northern North Atlantic is appropriate, even with complete NADW suppression. Clearly the model is incapable of producing the reported 15°C cooling over Greenland ((Johnsen et al., 1995; Schwander et al., 1997; Severinghaus, 1998). Our cooling of up to 7°C after 100 years of NADW cessation with the 32× St. Lawrence input (Fig. 8) is about twice that found by Manabe and Stouffer (1997), who however did not have a complete NADW shutdown (nor were they adding water at 0°C). It is inconsistent with the results of Schiller et al. (1997), who did find a cooling of 15°C in the vicinity of Greenland resulting from ≈0.4 Sv of freshwater input after several hundred years. The Schiller et al. response was associated with a reduction in poleward heat transport of about 0.6 PW, while in this model the reduction was 0.4 PW (Figure 9). Many differences exist between these two models, including the use of flux corrections, the assumption of large-scale geostrophic flow, and the "periodically synchronous" coupling of ocean and atmosphere in the Schiller et al. study; which if any of these factors accounts for this difference is uncertain. To some extent it is a geographic response issue rather than magnitude alone: it is shown in Part II that cooling of 10–15°C does occur in our model in the Arctic northeast of Scandinavia with NADW shutdown for some 200 years (where Schiller et al. find 20°C cooling) although the cooling near Greenland (3–4°C) is no greater than after 60 years of NADW cessation. Similarly, cooling of 6°C reported for the western Mediterranean (Cacho et al., 1999) does not arise in this model even with complete NADW cessation. The cooling magnitude is restrained by the maintenance of 50% of the poleward ocean heat transport despite NADW shutdown. This occurs primarily via the wind driven circulation, consistent with some estimates of its effect for the current climate (e.g., Covey and Barron, 1988); there is no evidence for a compensating increase in intermediate water production, and hence maintenance of strong ocean heat transports due to the thermohaline circulation.

Observations imply some cooling may have occurred over southern South America and southern Africa, with no effect over southeastern Australia (e.g., Rind et al., 1986). Even by years 41–60 in the 32× St. Lawrence run there is evidence of this (Fig. 6), and in these locales the cooling does not become more obvious later in the run. While tropical cooling was in general small at best over the oceans, some terrestrial locations did experience colder temperatures; strong cooling does arise in the more extreme experiments in the area of Huascaran (9°S, 77°W) (Fig. 6), in qualitative agreement with the ice core observations of Thompson et al. (1995). Some reductions in rainfall occurred in the Northern Hemisphere subtropics (Fig. 3), impinging upon the Cariaco Basin where drying was noted (Peterson et al., 2000). This is the result of increased subsidence associated with an intensification of the Hadley Circulation driven by warming south of the equator, which in turn results from reduced oceanic upwelling (as discussed in Paper II).

C. Implications for freshwater input

Simulations of the climate for the coming century tend to indicate reduced NADW production associated with freshening of the North Atlantic. In contrast to the experiments here, that freshening is initiated by increased precipitation in the North Atlantic, as tropical temperatures warm and more moisture is advected poleward (IPCC, 1995; Russell and Rind, 1999). It is then supplemented by reduced evaporation from the cooler ocean waters, as occurred here. In the experiment reported by Russell and Rind (1999), in a region in the North Atlantic of area about 20x106 km^2, precipitation minus evaporation increased over a 70- yr time period by about 0.3 mm d$^-$1, for a freshwater flux of about 0.07 Sv. Over the 70- year time frame that amounts to 4.9 Sv-yrs. Using the results for Table 3c for WEAK (with which that experiment was run), we estimate that the NADW reduction should have been (4.9) x (-4.63%), or about 23%. The actual reduction was about 30%, While obviously only approximate, the result does suggest that freshwater forcing of whatever nature in the North Atlantic (poleward of 55°N) has a similar effect on the coupled model's NADW response. Input of freshwater from other locations during the Pleistocene, from Greenland or western

Europe for example, might therefore also result in a similar sensitivity.

V. CONCLUSIONS

The primary conclusions of this study are as follows:

1. NADW production decreases linearly with the volume of freshwater added through the St. Lawrence. With a weaker initial deep water production, the decrease is proportionately larger.

2. With realistic input of freshwater for the Younger Dryas time period, it would take some 350 years for NADW to be completely shutdown. With a weaker initial circulation, it would take 150 years.

3. When freshwater input is ended after NADW shutdown, the circulation does not return for at least several decades (and as will be shown in Part II, for at least several centuries).

4. The cooling associated with freshwater input and NADW shutdown is not rapid and occurs over the entire Northern Hemisphere, in contrast to observations. With a smaller reduction in NADW production, the cooling over Greenland and tropical/Southern Hemisphere locations is too small.

5. NADW reinvigoration and rapid warming does not arise when freshwater input is ended.

6. The model sensitivity displayed to freshwater input through the St. Lawrence is similar to that associated with freshening due to the warming climate of the next century.

Acknowledgments. This research was supported by NSF grants ATM 96–1268, and OCE–99–05038. We thank Dorothy Peteet and Gerard Bond and Konrad Hughen for useful discussions, Tony Broccoli for his very helpful comments in review, and Jean Lerner for graphics assistance.

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D. Rind, G. L. Russell, S. Sheth, G. A. Schmidt, NASA Goddard Institute for Space Studies, 2880 Broadway, New York, NY 10025 (cddhr@giss.nasa.gov)

P. Dememocal, LDEO, Route 9w, Palisades, NY 10964

J. Teller, Department of Geological Sciences, University of Manitoba, Winnipeg, Manitoba R3T 2N2

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[1]NASA Goddard Institute for Space Studies and Center for Climate Systems Research, Columbia University, New York

[2]Lamont-Doherty Earth Observatory, Palisades, NY

[3]Now at Los Alamos National Laboratory, NM

[4]Department of Geological Sciences, University of Manitoba, Winnipeg, Manitoba

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This preprint was prepared with AGU's LATEX macros v5.01. File paper1 formatted May 10, 2001.

Table 1. Models and experiments used in this study.

NAME	DESCRIPTION
STAND	"Standard" coupled atmosphere-ocean model, $4{\times}5{\times}13$ levels, includes KPP mixing scheme
STAND (EXT)	Simulations beginning after 100 years of STAND with some changes to the model code.
WEAK	Coupled atmosphere-ocean model without KPP mixing scheme
FLUX	WEAK with salinity flux correction
Control Runs	Normal amount of freshwater added through the St. Lawrence (and other rivers) in STAND, WEAK, and FLUX
2×,4×,8× St. Lawrence	Additional freshwater added through the St. Lawrence, averaging 0.03 Sv, 0.06 Sv, 0.12 Sv, respectively (with STAND, WEAK and FLUX)
32× St. Lawrence	Additional freshwater averaging 0.53 Sv added through St. Lawrence in STAND
LTC(1)	"Realistic" freshwater input through the St. Lawrence and elsewhere, from Licciardi et al., 1999. Additional amount through the St. Lawrence of about 0.15 Sv, with an equivalent additional amount added through other source regions (e.g., Hudson Strait). Uses STAND.
LTC(2)	As in LTC(1) but uses STAND (EXT)

Table 2. Percentage (%) change in NADW stream function at 52°N, 900m in the different experiments as a function of time. Also shown are the average freshwater fluxes (Sv) through the St. Lawrence.

FRESHWATER	YEAR	6–10	21–25	46–50	71–75	96–100	116–120
STAND							
0.03	2× St Law	-3.8%	-2.5%	2.3%	4.5%	-5.4%	-11%
0.06	4× St Law	2.7%	0.6%	-4.4%	-12.23%	-9.8%	-20.6%
0.12	8× St Law	-2.8%	-11.6%	-12.2%	-15.7%	-32.3%	-44%
0.53	32×StLaw	-9.5%	-36.3%	-68.26%	-89.5%	-94.9%	-104.3%
0.15	LTC (1)	-4.1%	-6.8%	-7.6%	-16.3%	-16.3%	-28%
0.15	LTC (2)	4.9%	12.15%	-4.3%	-22.66%	-26.22%	-58.4%
FLUX							
0.03	2× St Law	-4.8%	-16.5%	-17.8%	-6.1%	-15.7%	-20%
0.06	4× St Law	5%	-19.6%	-27%	-14.6%	-23.9%	-24.5%
0.12	8× St Law	-9.35%	-23.6%	-39%	-38.6%	-51.1%	-51%
WEAK							
0.025	2× St Law	-1.4%	-8.1%	3.6%	2.9%	7.9%	-15.1%
0.055	4× St Law	-5.8%	-6.4%	-11.8%	2.0%	-13.5%	-54.7%
0.11	8× St Law	1.2%	-0.7%	-21.5%	-42.4%	-47.6%	-64.7%

Table 3a. Relationship between percentage change in N. Atl. stream function at 52°N (y) and St. Lawrence input in Sv (x), as $y = A(x)$, for different time periods. The linear correlation coefficient is also shown.

	STAND		WEAK		FLUX	
YEAR	A	r	A	r	A	r
0 to (6–10)	-17	0.81	11	0.16	-98	0.87
0 to (21–25)	-79	0.98	82	0.98	-77	0.98
0 to (46–50)	-137	0.998	-240	0.98	-203	0.99
0 to (71–75)	-167	0.99	-509	0.90	-298	0.999
0 to (96–100)	-177	0.99	-636	0.999	-368	0.99
0 to (116–120)	-167	0.98	-304	0.99	-296	0.997
AVE	–129	0.82	-283	0.64	-245	0.75

Table 3b. Relationship between percentage change in N. Atl. Stream function at 52°N (y) as a function of duration of fresh water input in years (t), as $y = B(t)$, for different St. Lawrence outflows. The linear correlation coefficient is also shown.

	STAND		WEAK		FLUX	
Sv	B	r	B	r	B	r
0.03*	-0.05	0.39	-0.02	0.11	-0.07	0.48
0.06	-0.19	0.95	-0.31	0.66	-0.17	0.63
0.10	-0.34	0.95	-0.61	0.99	-0.36	0.94
0.50	-0.83	0.96				
AVE	-0.35	0.43	-0.32	0.58	-0.20	0.53

*Except for WEAK, where the value is 0.017

Table 3c. Relationship between percentage change in N. Atl. Stream function at 52°N (y) as a function of volume of fresh water input (v) (in Sv- years), as $y = C(v)$ for the different runs. Results are for the simulations after 120 years. The linear correlation coefficient is also shown, as is the volume flux necessary to produce 100% reduction in NADW production as calculated from the equation.

	STAND		WEAK		FLUX	
	C	r	C	r	C	r
120 yrs	-1.80	0.965	-4.63	0.892	-3.12	0.867
100% reduction (Sv yrs)	55.5		21.6		32	

Table 4. Change in parameters influencing the temperature and radiation between the 32× St. Lawrence input experiment and the control for years 141–160.

	GLOBAL	NH	SH
Surface Air Temp. (°C)	-1.12	-1.92	-0.33
Atmos. Air Temp. (°C)	-0.80	-1.10	-0.40
Atmos. Water Vapor (%)	-4.5	-8.0	-0.8
Sea Ice Cover (%, abs)	1.24	2.04	0.45
Surf. Albedo (%, abs)	0.69	1.18	0.16
Cloud Cover (%, abs)	-0.60	-1.01	-0.18
Planetary Albedo (%, abs)	0.34	0.52	0.17
Net Rad. TOA (Wm^{-2})	0.29	-0.02	0.59

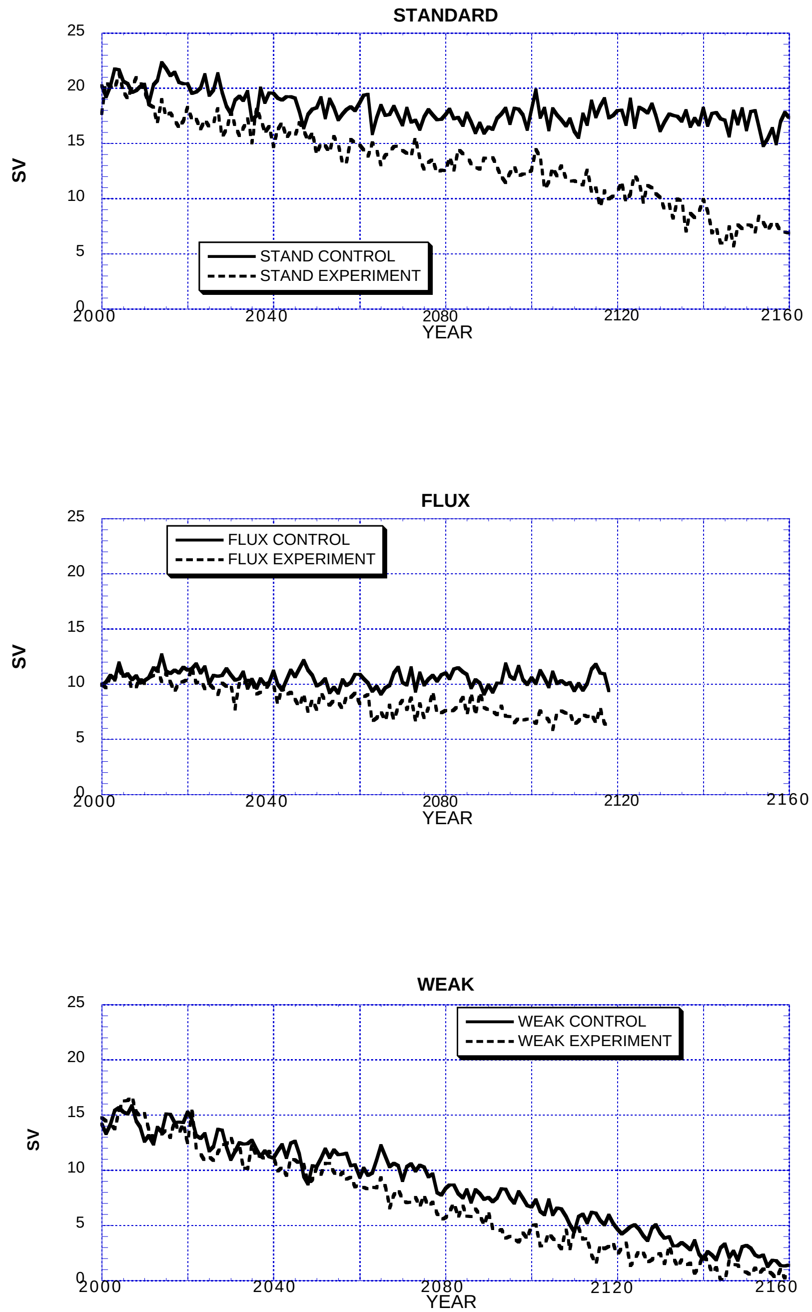

Figure 1. North Atlantic mass streamfunction (Sv) at ≈50°N, 900m in STAND (top), FLUX (middle) and WEAK (bottom) for the control run and 8×St. Lawrence experiments (inflow of 0.12 Sv). Note that this inflow is similar to that estimated through the St. Lawrence for the Younger Dryas deglaciation.

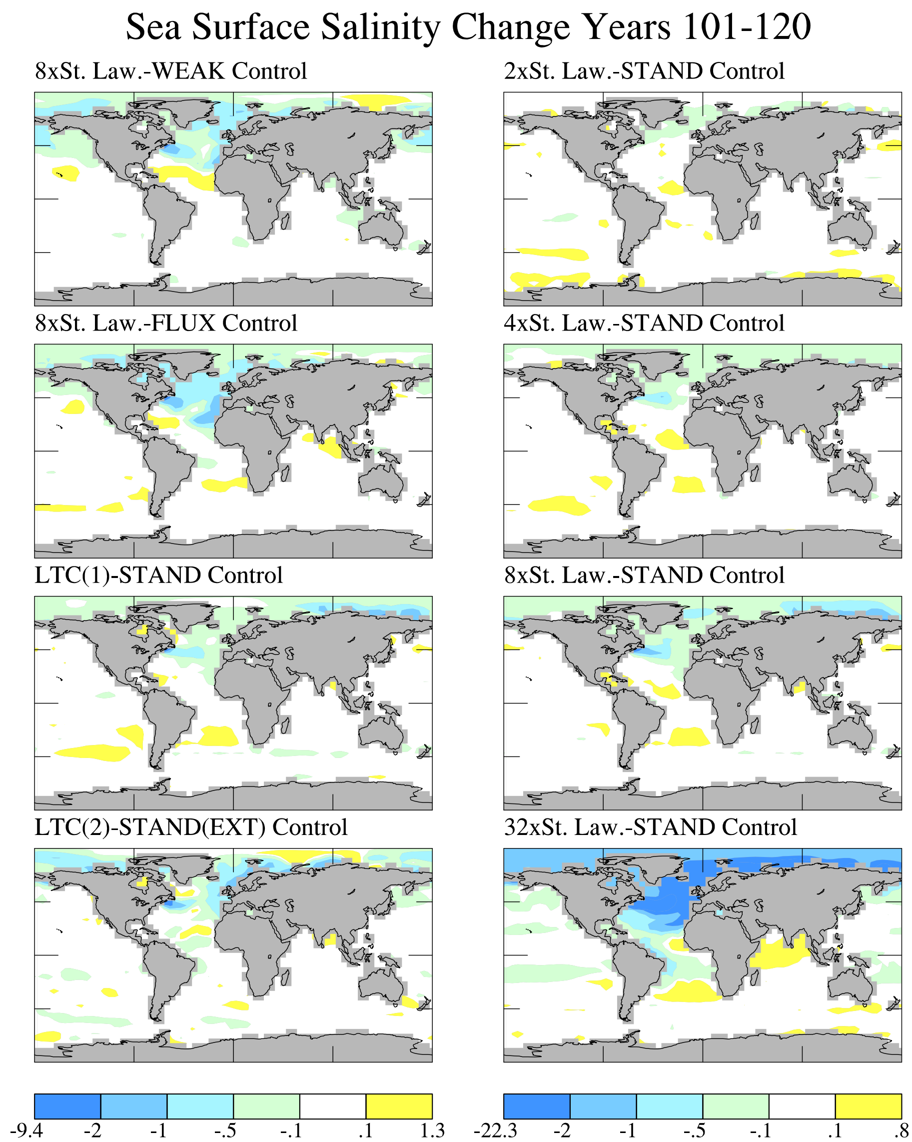

Figure 2. Sea surface salinity change during years 101 through 120 in different experiments.

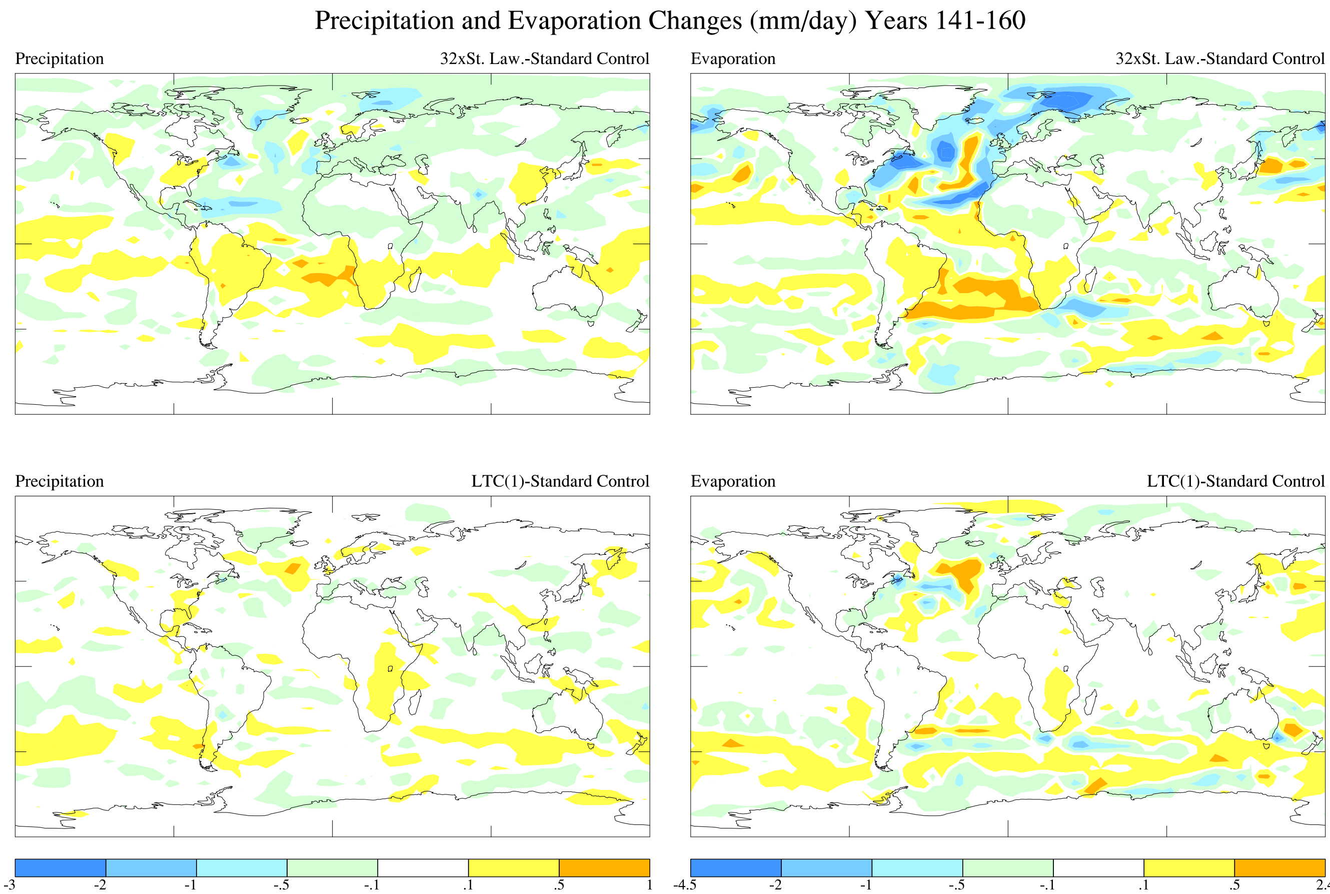

Figure 3. Precipitation (left) and evaporation (right) change during years 141–160 in the 32× St. Law. run (top) and LTC(1) (bottom).

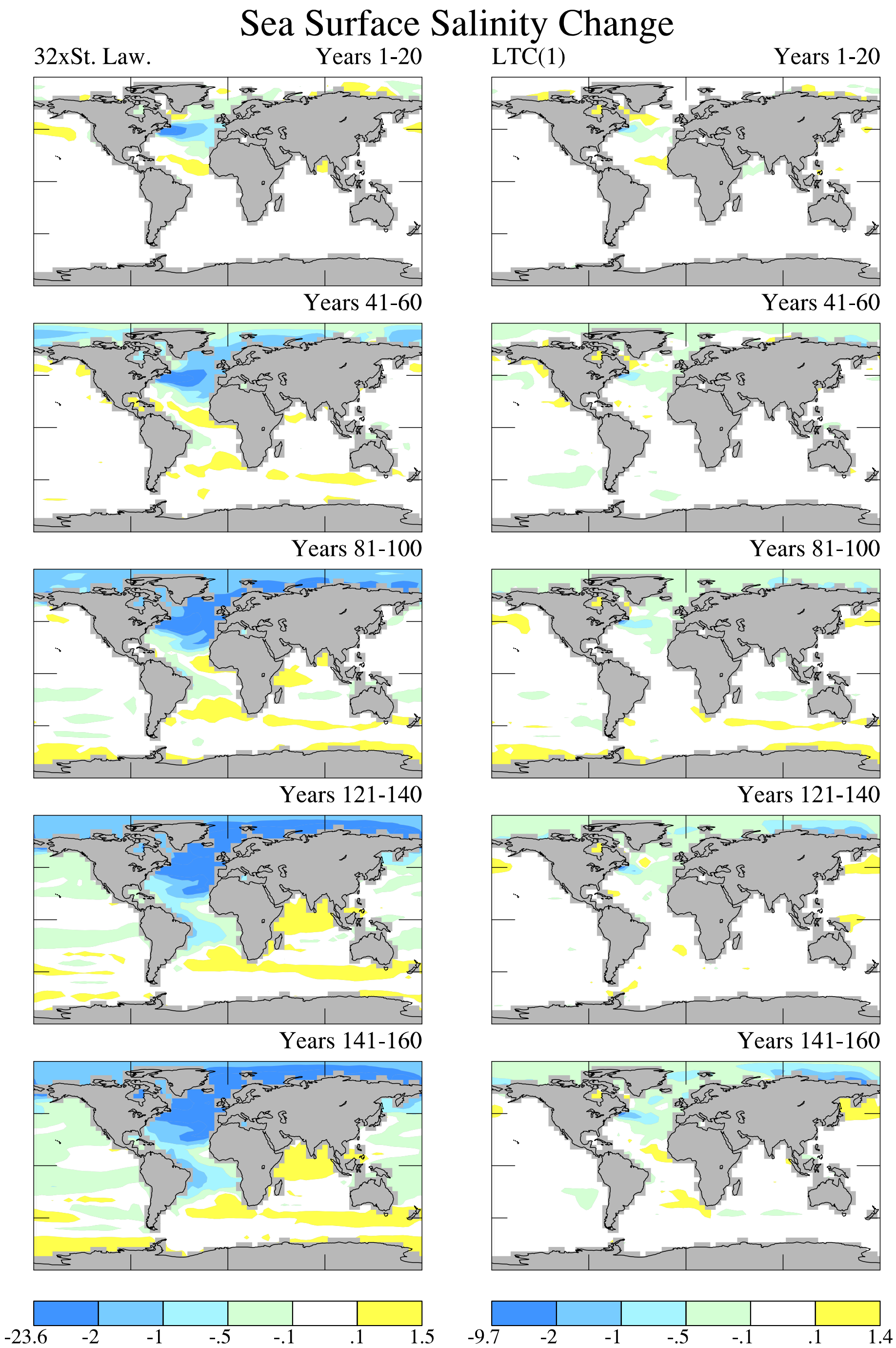

Figure 4. Sea surface salinity change with time in the 32× St. Law. run (left) and LTC(1) (right).

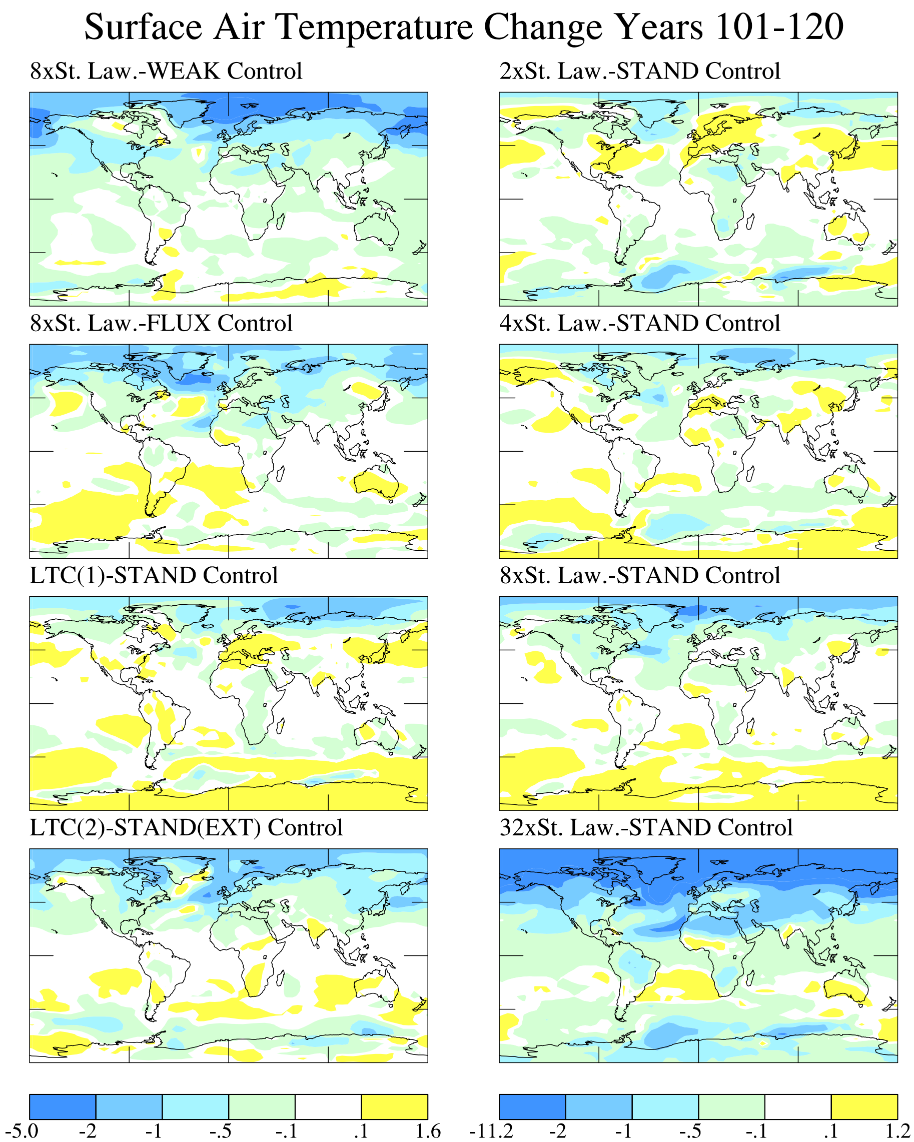

Figure 5. Surface air temperature change during years 101 through 120 in different experiments.

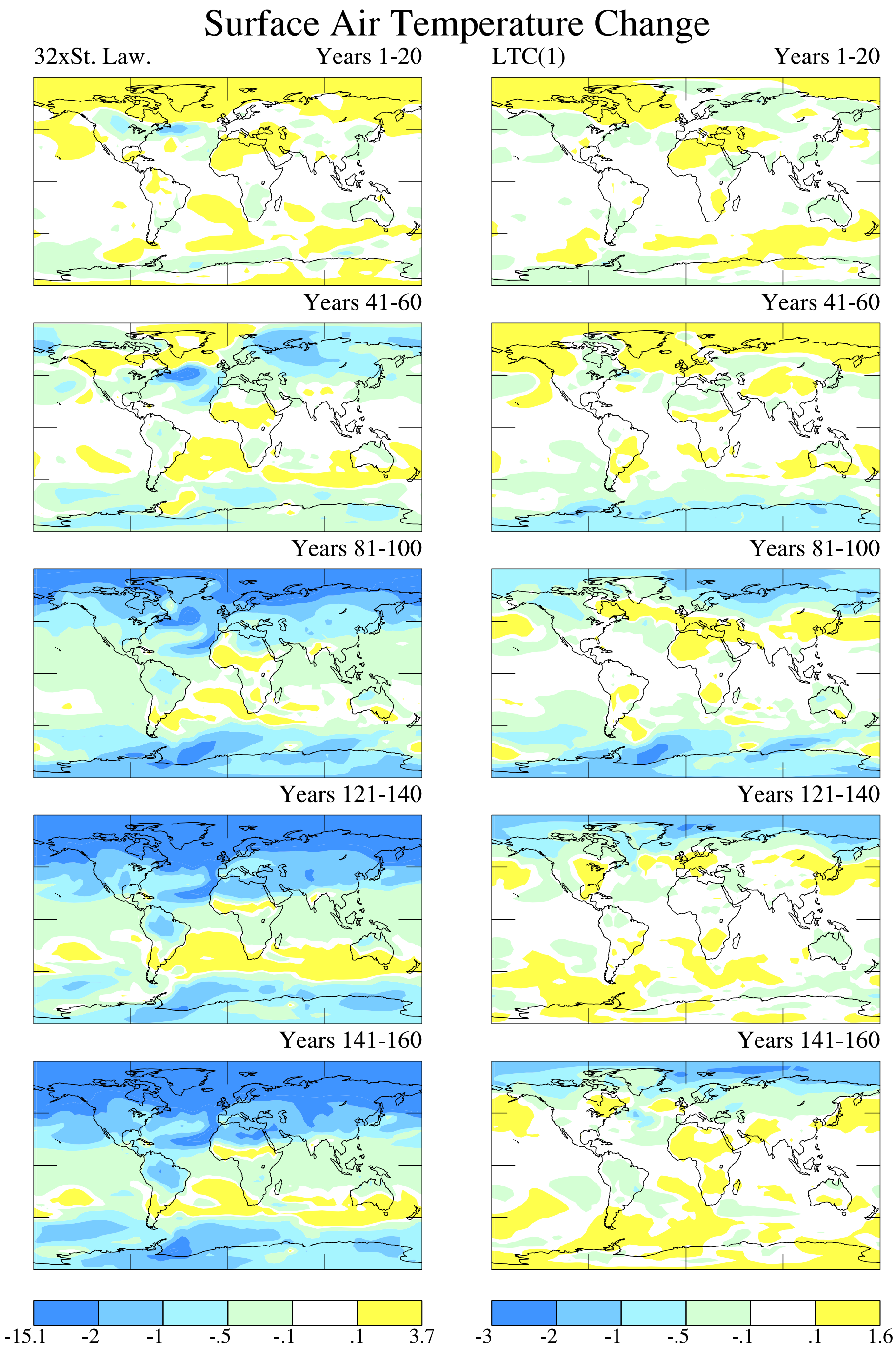

Figure 6. Surface air temperature change with time in the 32× St. Law. run (left) and LTC(1) (right).

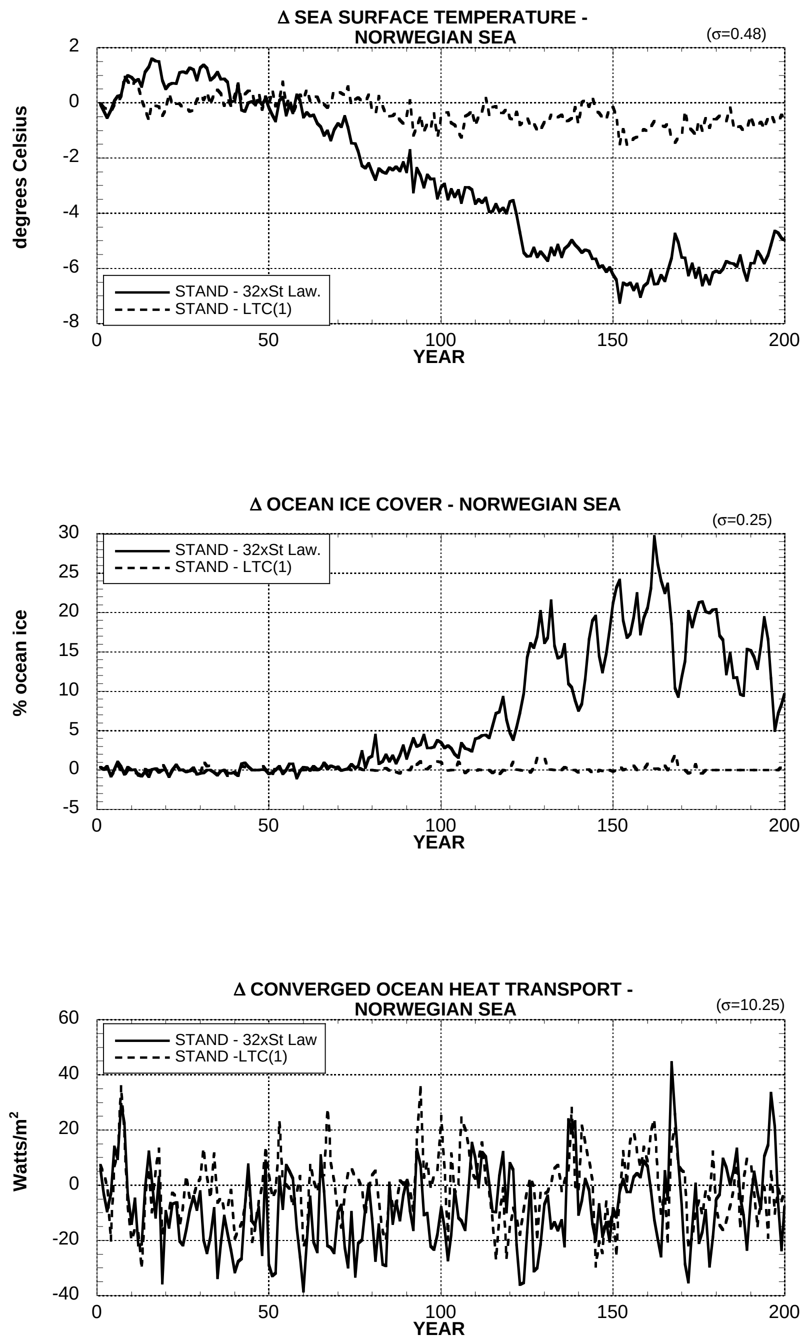

Figure 7. Sea surface temperature change (top), change in ocean ice coverage (middle), and change in converged ocean heat transport (bottom) for the Norwegian Sea as a function of time in the 32× St. Law. and LTC(1) experiments. Note that freshwater input is ended in year 160.

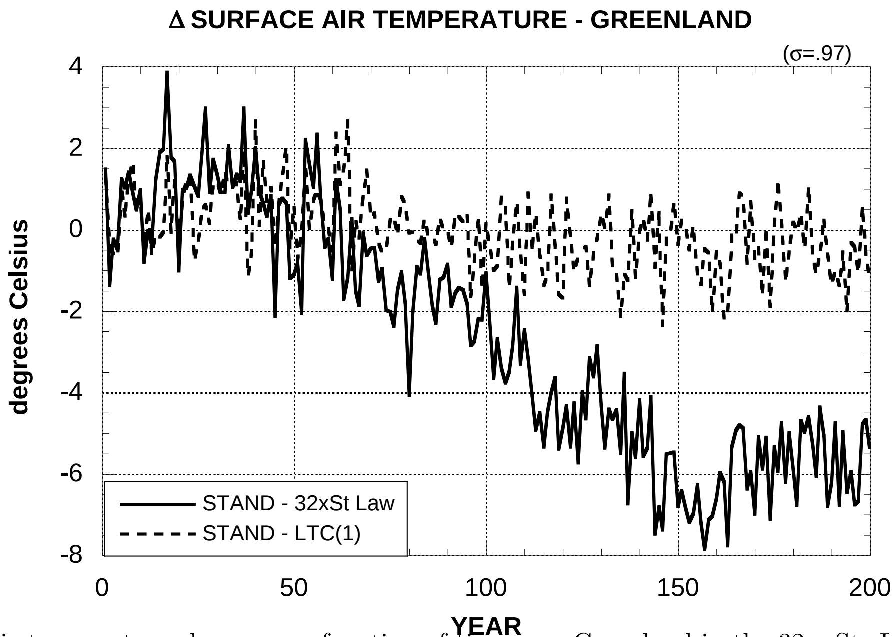

Figure 8. Surface air temperature change as a function of time over Greenland in the 32× St. Law. and LTC(1) experiments.

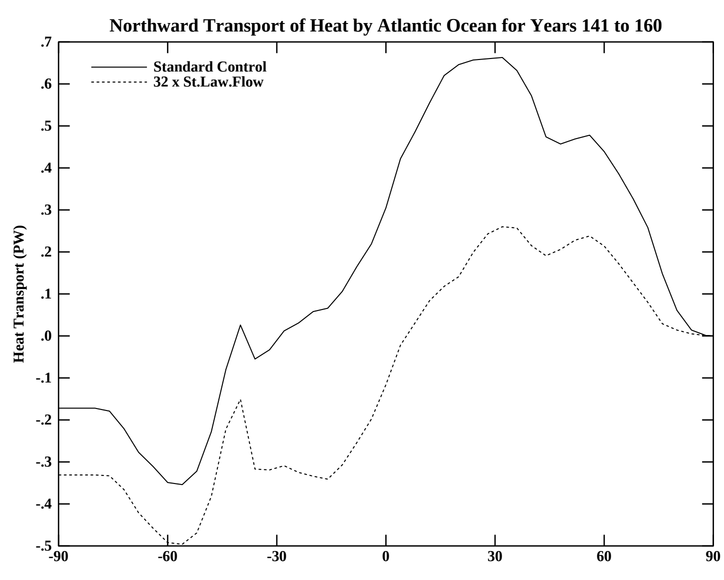

Figure 9. Northward energy transport in the oceans (PW) as a function of latitude in years 141–160 in the 32× St. Law and STAND control.

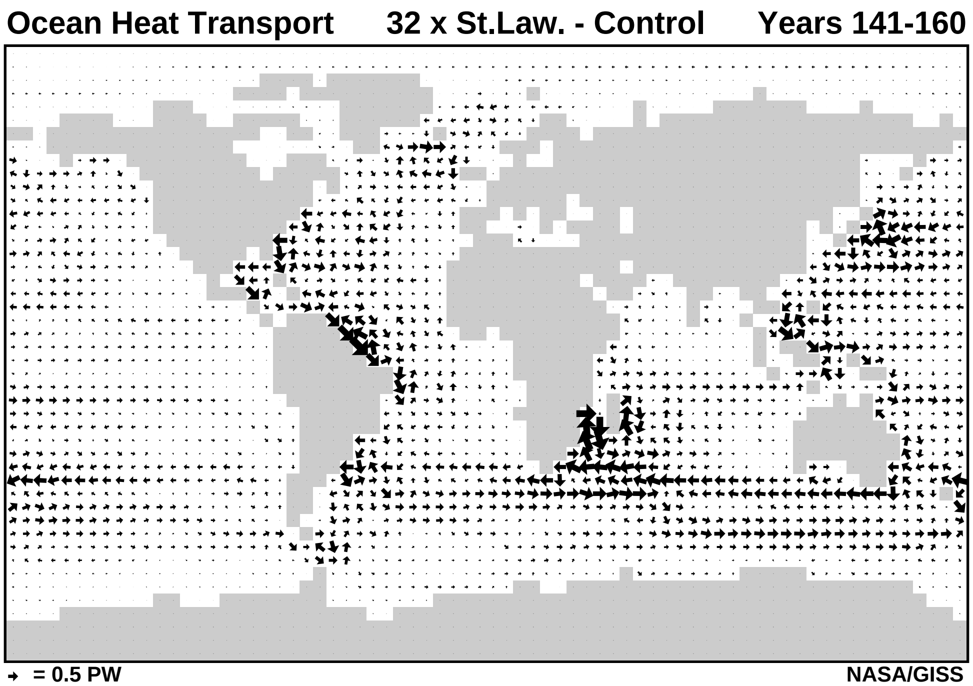

Figure 10. Vertically-integrated change in potential enthalpy flux (PW) between the 32× St. Law. experiment and the control for years 141–160.

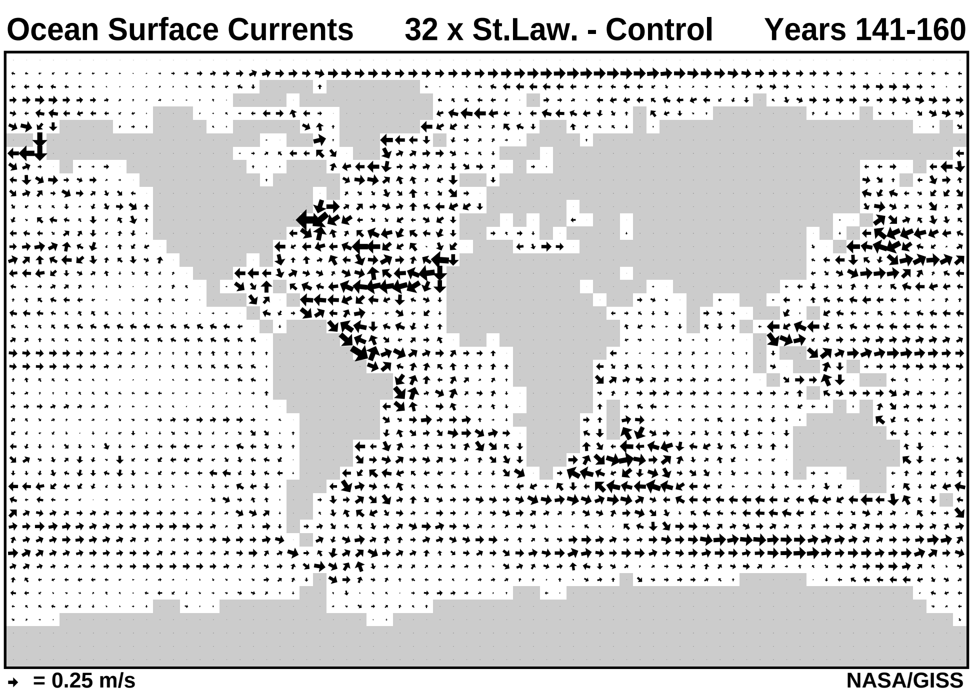

Figure 11. As in Fig. 10 except for the change in the ocean currents (m/s) in the first model layer.